Non-adiabaticity in mantle convection

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Abstract

Seismic observations indicate that Earth's lower mantle is homogeneous as revealed by smooth depth variations of bulk sound velocity and Bullen's inhomogeneity parameter being close to one. Here we show with 3D spherical convection simulations that it should also be non-adiabatic because a significant proportion of mantle heat sources is internal. The computer simulations predict nearly isothermal lower mantle geotherms with values of the seismologically observable inhomogeneity parameter $\eta \sim 1.01$, implying that the temperature drop across the Core Mantle Boundary exceeds 1000 degrees.

1. Introduction

Seismic studies of body waves and free oscillations provide by far the most detailed information on the large scale structure of the Earth. Throughout most of the lower mantle they reveal smooth depth variations of seismic sound velocities, consistent with a homogeneous material under adiabatic self compression. Assuming adiabaticity *Birch* [1952] combined the Adams-Williamson equation with bulk sound velocity measurements to calculate a lower mantle density profile. Lower mantle density can also be obtained directly, with no assumption of adiabaticity, from the study of spheroidal low frequency normal modes. The agreement between body wave and normal data suggests that the lower mantle is indeed nearly adiabatic [*Masters*, 1979; *Dziewonski & Anderson*, 1981] and modeled geotherms have been found in good agreement with the adiabat [*Shankland & Brown*, 1985], supporting a long-held view that vigorous convection of the deep mantle results in adiabaticity.

The apparent lower mantle adiabaticity is a problem in geodynamics, because geotherms derived from fluid mantle models are rarely adiabatic. They are characterized instead by nearly isothermal temperatures in the interior of the mantle with strongly superadiabatic temperature gradients concentrated in the thermal boundary layers near the surface and the core mantle boundary (CMB), even as the Rayleigh number approaches the convective vigor of the mantle [*Tackley et al.*, 1994; *Parmentier et al.*, 1994; *Bunge et al.*, 1997]. A variety of mechanisms attempts to explain this paradox: (i) the effects of "mantle avalanches" [*Tackley et al.*, 1994], (ii) insufficient vigor of mantle convection models [*Parmentier et al.*, 1994], and (iii) predominantly internal mantle heat generation [*Jeanloz & Morris*, 1987].

The latter mechanism is supported by observational evidence, which places a lower bound of about 10% core heating on the overall heating budget of the mantle [*Davies*, 1988]. The physical reason for subadiabaticity is that material is slowly reheated by radioactivity while it is entrained toward the surface by diffuse upwellings. *Jeanloz & Morris* [1987] estimated the effect of internal heating on the mantle geotherm, and here we further investigate it in fully dynamic models, demonstrating that subadiabaticity of 100-300 degrees may be expected in most of the lower mantle, while superadiabaticity in the bottom boundary layer should be confined to a layer of $\sim$ 500 km.

2. Convection models

We model 3D spherical convection using the code TERRA [*Bunge & Baumgardner*, 1995]. Figure 1a shows an isoviscous reference calculation obtained, assuming the mantle is compressible and heated purely from within. Material properties are identical to *Bunge et al.* [1997], and the Rayleigh number is 1.1 X 10^8 based on internal heating. The near surface and interior (cut-away) temperature fields are shown and the reference adiabat has been subtracted. Temperatures are subadiabatic as evidenced by colder fluid being concentrated at the mantle base. Shown in Figure 1e is the non-adiabatic radial temperature profile. Below the superadiabatic upper thermal boundary layer, temperatures drop subadiabatically by about 300 degrees from the upper mantle to the CMB. Figure 1b shows a model identical to the reference case, except that we have stiffened the lower mantle by a factor of 40 relative to the upper mantle viscosity (2 X 10^{22} Pa s), as suggested by studies of the geoid [*Ricard et al.*, 1993].

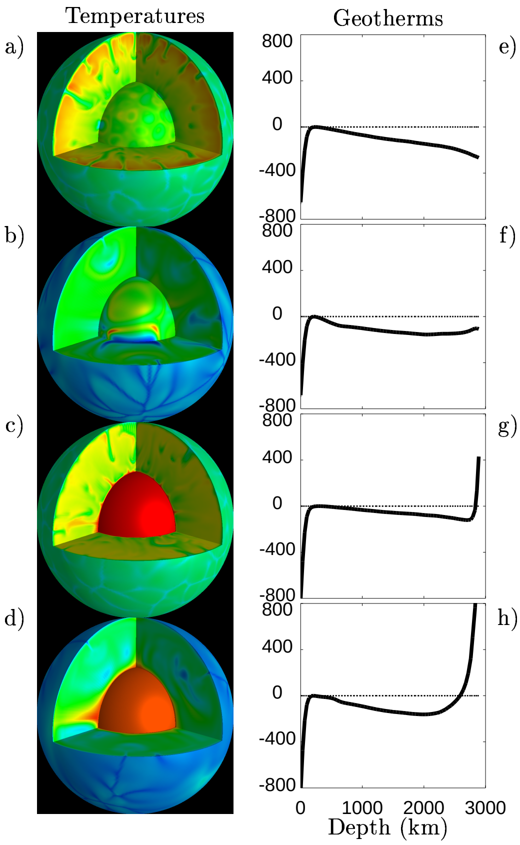

Figure 1. a) Non-adiabatic temperatures for the isoviscous, purely internally heated reference case. Blue is cold and red is hot. The top 100 km of the mantle is removed. Cold fluid is concentrated near the mantle base. b) Same as (a) except that lower mantle viscosity has been increased by a factor of 40. c/d) Same as (a/b) except for the addition of 50% core heating. (e)-(h) Non-adiabatic geotherms for cases (a)-(d). Superadiabaticity is concentrated into narrow thermal boundary layers. The geotherm is subadiabatic for intervening regions.

Subadiabaticity (Figure 1f) is comparable to the reference case, except for a region near the CMB where temperatures increase somewhat steeper than the adiabat. An isoviscous convection model with 50% core heating is shown in Figure 1c. Note that this core heat flux is significantly larger than estimates for the Earth [*Davies* 1988], and that it is in addition to the internal heating rate, which remains unchanged from the reference case. The temperature field shows numerous hot upwellings from the CMB in addition to cold downwellings from the surface. The radial temperature profile (Figure 1g) reveals that mantle subadiabaticity is reduced to half its value in the reference case, or about 150 degrees. A model combining a lower mantle viscosity increase (factor 40) and 50% core heating with the internal heat production of the reference calculation is shown in Figure 1d. In this hybrid model hot mantle upwellings from the CMB are swept into a few major plumes due the long-wavelength planform induced by viscosity layering [*Bunge et al.*, 1996], while most of the lower mantle away from plumes appears to be relatively cold. This observation is borne out in the radial temperature profile (Figure 1f), which shows a non-adiabatic temperature gradient of -0.1 degree/km before reaching a broad thermal boundary layer at the CMB.

3. Density profiles and seismic inhomogeneity

The modeled geotherms may be converted into density profiles to estimate the effect of subadiabaticity on mantle density. In Figure 2a we calculate the lower mantle excess density relative to the adiabatic reference density profile in a mineralogic model of pyrolite composition composition [*Matas*, 1999]. As expected, the cold geotherms steepen the density increase with depth for all four models shown in Figure 1. However, the anomalous excess density remains

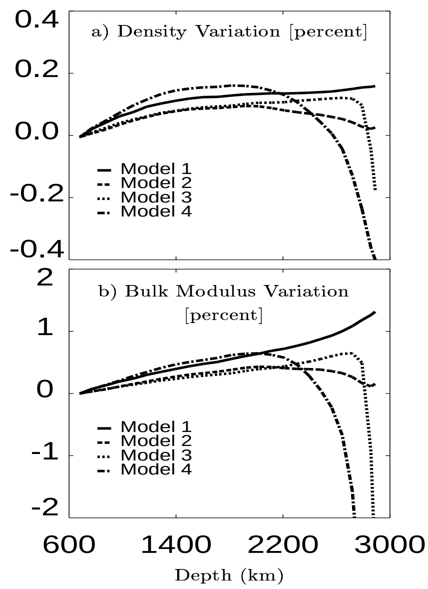

Figure 2. a) Lower mantle excess density in a pyrolite model relative to the adiabat for the four geotherms in Figure 1. Density variations remain below 0.2%. b) Same as a), but for the isentropic bulk modus K_s.

below 0.2% due to the small value of the thermal expansivity under lower mantle conditions (1.5 X 10^{-5} 1/K) [*Chopelas & Boehler*, 1992; *Duffy & Ahrens*, 1993]. The effect of subadiabaticity on the isentropic bulk modulus K_s (Figure 2b) is somewhat larger, with excess values of K_s approaching one percent. This one-percent increase brings K_s close to the isothermal value of the bulk modulus, implying a somewhat stiffer compressional response of pyrolite for the modeled geotherms. To go one step further in comparing our results with seismic observations, we compute Bullen's stratification parameter η [*Bullen*, 1963] in Figure 3. η measures departures from the Adams-Williamson relation, and is exactly one for adiabatic and chemically homogeneous regions. From Figure 3 we identify three distinct regimes for η. In the thermal boundary layers near the top and bottom of the mantle η is substantially smaller than one, as expected. There is also a sharp increase in the upper mantle for models with layered viscosity due to the strongly subadiabatic geotherm in the low viscosity zone. For the rapidly convecting interior of the lower mantle, we find η is about 1.01.

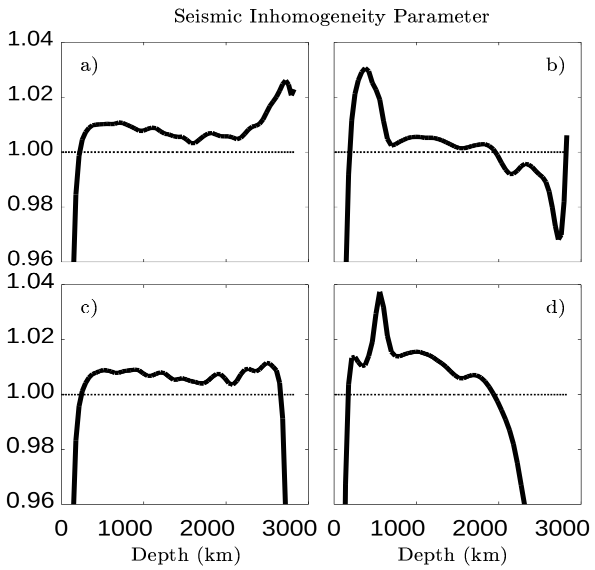

Figure 3. (a)-(d) Seismic inhomogeneity parameter η (see text) for the four models in Fig. 1a-d. η is defined as $\eta = -\frac{\phi}{\rho g}\frac{d\rho}{dr}$, where ρ is density, r is radial distance from the Earth's center, g is the acceleration due to gravity, and ϕ $(= \frac{K_s}{\rho})$ is the seismic parameter.

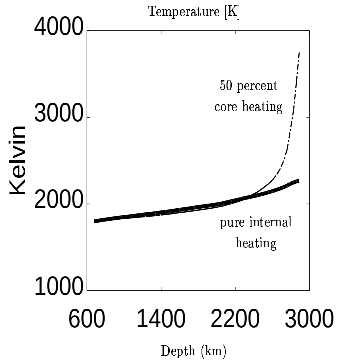

Figure 4. Lower mantle geotherms for the isoviscous, bottom insulated reference model (Fig. 1a), and the stratified viscosity case with 50% heating from the core (Fig. 1d). The geotherms are nearly isothermal away from the thermal boundary layer at the CMB.

4. Discussion

The magnitude of mantle non-adiabaticity implied by the convection models is significant. In most of the lower mantle subadiabaticity is comparable (~1/3) to the temperature change across a thermal boundary layer, the largest perturbation of the geotherm resulting from convection. Observational evidence [*Davies*, 1988] suggests predominantly internal heat generation for the mantle, with an additional component arising possibly from secular cooling [*Davies*, 1980], which would act as a further source of internal heat. We explore the consequences in Figure 4, where we show absolute lower mantle temperatures tied to a temperature of 1800 K at the 660 km depth spinel/post-spinel phase change for two end-member calculations. The geotherm computed from the isoviscous, purely internally heated reference model (Figure 1a) is nearly isothermal, because subadiabaticity nearly cancels the adiabatic temperature rise with depth. The mantle temperature inferred for the CMB in this bottom insulated case is 2200 K. Setting the core temperature to a geophysically more plausible value (3800 K) [*Boehler*, 1996] results in 50% heating from the core even for the more realistic convection model with a high viscosity lower mantle (Figure 1d), due to the sharp temperature drop of nearly 1600 K across the thermal boundary layer. This modeled core heat flux is substantially higher than the core heat flux inferred from mantle hotspots [*Davies*, 1988]. A chemically distinct layer at the mantle base would lower the core heat flux by providing an additional thermal boundary layer, and would also help to reduce the high excess temperature of mantle plumes [*Farnetani*, 1997] suggested by our calculations. The existence of this layer appears likely from geochemical [*Hofmann*, 1997] and seismic considerations [*van der Hilst & Karason*, 1999], and has been explored in dynamic models [*Christensen & Hofmann*, 1994; *Kellogg et al.*, 1999; *Coltice & Ricard*, 2000]. A seismological consequence of our result is that homogeneous regions in the lower mantle should appear otherwise, with η ~1.01, and that η should be as low as η ~0.92 for a broad thermal boundary layer region in the bottom 500 km of the mantle (Figure 3). Seismic observations of η are resolved to about 2% in the lower mantle [*Masters*, 1979], which suggests that constraints on η should be improved to infer the mantle geotherm.

Acknowledgments.

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